Annual Technical Report
99HQGR0035

Title: Radiated Energy and State of Stress During Earthquake Rupture

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Abstract

We determined the energy release in earthquakes using the data obtained from TriNet in southern California. We investigated the effects of the depth, the mechanism and the propagation path by examining the results for events with different mechanisms and depths. We found that the path and site effects dominate so much that these effects are not obvious. By far the most dominant is the path-site effects. We determined the station site corrections for the integral of the square of ground motion velocity. The amplification factors are very large: a factor of 10 is common, and for some stations, it is as large as 30. Source directivity can produce a strong azimuthal variation of energy radiation. A good example is the 1992 Landers earthquake in which strong directivity was observed over the frequency band of energy spectrum. We developed a numerical method to correct for this effect. We found that the energy estimate obtained for the Landers earthquake listed in K93 is overestimated by a factor of 2.6. With these corrections, we determined the energy released by earthquakes which occurred in southern California for the period of 1995 to the present. Also, we updated the results for the larger earthquakes (1991 Sierra Madre earthquakes (M=5.8), 1992 Joshua Tree earthquake (*M*=6.4), 1992 Landers earthquake (*M*=7.3), 1992 Big Bear earthquake (*M*=6.4), and the 1994 Northridge earthquake (*M*=6.7)) applying the newly determined station corrections and directivity effects. The results show that the energy to seismic moment ratio, $\tilde{e}$, for large earthquakes is 10 to 100 times larger than for small earthquakes. This can be interpreted in term of reduction of friction when the slip exceeds a threshold value, about 10 to 50 cm, for large earthquakes.

Introduction

How the stress on a fault plane changes as a function of slip during dynamic rupture is a fundamentally important problem for understanding the fault constitutive relations [e.g. Scholz, 1990] and the wave form of ground motion, especially in the near-field of a large earthquake [e.g. Heaton, 1990; Heaton et al., 1995]. To this end, many investigations have been made to determine the slip function on a fault plane by inversion of observed seismograms. Two difficulties have been encountered in this approach. First, in most modeling studies, short-period (usually 2 sec or shorter) waves are filtered out because of the difficulty in modeling such short-period waves. At periods shorter than 2 sec, scattering of waves and possible complexities of source process produce too complex wave forms to be modeled with a simple model. Second, the observed wave form is a convolution of both local slip function which is primarily controlled by the stress on the fault plane and the rupture function, and it is not always easy to separate the two functions. Nevertheless, many interesting results have been obtained from these studies, which contributed significantly to our understanding of earthquake physics [e.g. Quin, 1990; Miyatake, 1992; Wald and Heaton, 1994; Beroza and Mikumo, 1996; Bouchon, 1997; Ide and Takeo, 1997]

We investigated this problem from a different angle. Namely, we used an integrated quantity, radiated energy, to gain some insight into this problem. In this approach, we use the total energy contained in the radiated wave field, thereby avoiding the difficulties caused by filtering of short period waves in wave-form modeling studies. We have recently applied this method to determine the state of stress during the rupture of the 1994 deep Bolivia earthquake [Kanamori et al. 1998]. From the radiated energy, combined with the static stress drop, we could determine fairly definitively that the fracture energy involved was very large, which in turn suggests a gradual stress change during faulting. This example is encouraging that the quality of presently available seismic data is good enough to gain useful insight into the physical processes of dynamic rupture.

This approach, however, has some weakness. First, estimation of radiated energy is not as easy as it seems, because of the very complex wave propagation effects in Earth's crust. This difficulty has been demonstrated by the large discrepancy, sometimes a factor of 50, between the estimates obtained by different investigators [e.g. Singh and Ordaz, 1994]. Second, as will be discussed later, interpretation of the results in terms of stress is indirect and leaves some uncertainties in the final model. Nevertheless, in view of the importance of understanding the basic physics of earthquake rupture and its effects on near-field ground motion, we took this approach by improving the current methodology for energy estimation. With the availability of high-quality broad-band data, we could reduce the uncertainty of energy estimates.

Model

We briefly describe the basic concept, seismic energy budget, underlying this project .

The energy budget of earthquakes has been extensively studied by many investigators [e.g. Knopoff, 1958; Dahlen, 1977; Kostrov, 1974; Savage and Walsh; 1978]. Following these studies, and referring to Orowan [1960] and Savage and Wood [1971], here we consider a simple stress-release model. An earthquake is viewed as a stress release process on a surface *S*. At the initiation of an earthquake, the initial (before an earthquake) shear stress on the fault plane σ_0 drops to a constant dynamic friction σ_f. If the condition for instability is satisfied [Brace and Byerlee; 1966, Scholz, 1990], rapid fault slip motion begins and eventually stops. At the end, the stress on the fault plane is σ_1 (final stress) and the average slip (offset) is *D*. In this example $\sigma_f = \sigma_1$. The difference $\Delta\sigma_s = \sigma_0 - \sigma_1$ is the static stress drop, and the difference $\Delta\sigma_d = \sigma_0 - \sigma_f$ is the driving stress of fault motion and is usually called the dynamic stress drop or effective tectonic stress [Brune, 1970]. During this process, the potential energy (strain energy plus gravitational energy) of the system, *W*, drops to $W - \Delta W$ where ΔW is the strain energy drop, and seismic wave is radiated carrying energy E_R. Then the energy budget can be written as

$$\Delta W = E_R + E_F + E_G \tag{1}$$

where E_F is the frictional energy loss given by $E_F = \sigma_f DS$, and E_G is the fracture energy.

Knopoff [1957], Dahlen [1977] and Kostrov [1974] showed that $\Delta W = \bar{\sigma} DS$ where $\bar{\sigma} = (\sigma_0 + \sigma_1)/2$ is the average stress during faulting. From (1), we obtain

$$E_R = (\sigma_0 + \sigma_1)DS/2 - \sigma_f DS - E_G = (1/2)(2\Delta\sigma_d - \Delta\sigma_s)DS - E_G$$

$$= M_0(2\Delta\sigma_d - \Delta\sigma_s)/2\mu - E_G \tag{2}$$

where $M_0 = \mu DS$ is the seismic moment, and μ is the rigidity. This is a simple but fundamental relationship which does not involve major assumptions. The fracture energy E_G can be ignored for most shallow earthquakes, and (2) can be written as

$$E_R = M_0(2\Delta\sigma_d - \Delta\sigma_s)/2\mu \tag{2'}$$

A similar relation has been used in seismology [e.g. Savage and Wood, 1971], but this particular form introduced here is useful because E_R is expressed in terms of the specific physical parameters $\Delta\sigma_s$ and $\Delta\sigma_d$ which directly characterize the stress release process on the fault plane.

The variation of stress during faulting can be more complex For example, the stress may increase in the beginning of the slip motion because of loading caused by advancing rupture, or of a specific friction law such as the state-rate dependent friction law [Dieterich, 1979]. In fact, seismological inversion studies have shown this increase

[Quin, 1990; Miyatake, 1992; Mikumo and Miyatake, 1993; Beroza and Mikumo, 1996; Ide, 1997]. However, this increase is of short duration and the amount of slip is small so that little energy is radiated. Thus, we do not include it in our energy budget.

The friction may not be constant during faulting. It may drop drastically in the beginning and later resumes a somewhat larger value, or it may decrease gradually to a constant level. The latter is called a slip-weakening process. If the friction is not constant, the rupture dynamics is complicated, but for the energy budget considered here, we formulate this problem for a simple case. The friction σ_f gradually drops to a constant value σ_{f0} until the slip becomes D_c. In general, the final stress σ_1 can be different from σ_{f0}. Then, we define the average friction $\bar{\sigma}_f$ by

$$\bar{\sigma}_f = \frac{1}{D}\int_0^D \sigma_f(u)du \tag{3}$$

where u is the slip (offset) on the fault plane and D is the total offset. Then, equation (2') can be written as

$$E_R = M_0\left(2\Delta\sigma'_d - \Delta\sigma_s\right)/2\mu \tag{4}$$

where

$$\Delta\sigma'_d = \sigma_0 - \bar{\sigma}_f \tag{5}$$

Here, $\Delta\sigma'_d$ defined by (5) can be still called the dynamic stress drop, but it is slightly different from that traditionally used.

From seismic observations we determine the scaled energy, $\tilde{e} = E_R / M_0$, and $\Delta\sigma'_d$ (frome equation 4), from which we infer how the stress changed during faulting.

Results

The fundamental seismological parameters we determine are the radiated energy E_R and seismic moment M_0. The seismic moment can be determined accurately using the waveform inversion method, but the energy determination is still difficult.

Determination of Radiated Energy

The basic quantity involved in computation of energy is

$$f(\Delta,\ mechanism,\ depth,\ source\ \ spectrum) = \int_0^T v^2(\Delta,\ t)dt \tag{6}$$

where $v(\Delta)$ is the ground motion velocity at a distance Δ, and the integral is over the signal duration T. Since the observed seismogram includes P, S, and surface waves, and the partition of energy into these phases depends on the mechanism, depth, and the source spectrum, $f(\Delta)$ depends on these factors. Since propagation of short-period waves in Earth's crust is severely affected by all kinds of effects, e.g. attenuation, scattering, site response etc, it is difficult to treat this problem completely deterministically.

Hence, we used the semi-empirical method described in Kanamori et al. [1993, hereafter abbreviated as K93] for the events larger than M=3.7 recorded with TriNet. The recent results obtained by Mayeda et al. [1996] who used a completely different method agree with those of K93 within a factor of 2. When the method was developed in 1993, only a few digital stations were available, and the average distance to the stations used was fairly large, commonly larger than 150 km. Since the main cause of the errors is the propagation effect, the method works much better now, because the network is much more dense and the average distance to the stations is shorter.

We investigated the effects of the depth, the mechanism and the propagation path by examining the results for events with different mechanisms and depths. We found, somewhat surprisingly, that the path and site effects dominate so much that these effects are not obvious. By far the most dominant is the path-site effects. Thus we focussed our efforts on the determination of the site amplification factors.

The TriNet stations are located at sites with various site conditions. We have noticed that significant amplification occurs at many of the TriNet stations. This amplification effect is complex, and at present, we can remove it only by applying empirical station corrections. For energy estimation, station corrections for the integral of the square of ground motion velocity have been empirically determined, and as new stations are deployed, they are constantly updated. The station Pasadena is used as reference. The amplification factors are very large: a factor of 10 is common, and for some stations, it is as large as 30. In principle, station corrections must be a function of distance and azimuth, but so far the data are still insufficient to determine the distance-azimuth-dependent station corrections.

Source directivity can produce a strong azimuthal variation of energy radiation. If the azimuthal coverage of the station is dense, the azimuthal variation is usually averaged out, but, with a limited azimuthal coverage, this can be a problem. A good example is the 1992 Landers earthquake in which strong directivity was observed over the frequency band of energy spectrum. We developed a numerical method to correct for this effect. We found that the energy estimate obtained for the Landers earthquake listed in K93 is overestimated by a factor of 2.6, because many of the stations that recorded the ground motion of the Landers earthquake happened to be in the direction of maximum radiation.

The directivity effect should be present for smaller earthquakes too, but the effect is not obvious. For small events, the directivity effect is expected to show up at relatively short period. However, the short-period waves are scattered extensively, and the directivity pattern in radiation is obscured. Thus, the directivity effect for smaller earthquakes is not explicitly considered.

An example is a M=4.9 earthquake which occurred on April 27, 1997, in the Northridge area. The wave forms change very drastically reflecting the complex path effects and the site effects. Then the radiated energy is computed from each station using the method of K93, and the station corrections applied. Even after the application of station corrections, almost an order of magnitude variation in the energy estimate exists. However, the logarithmic standard deviation around the mean is about 0.4, suggesting that the energy estimate is accurate within a factor of 2 to 3. We performed a detailed source inversion for this event using the 4 close-in stations (CALB, NOT, OSI, and SOT), and could estimate the radiated energy independently using the method described by Vassiliou and Kanamori [1982]. The result agrees within a factor of 2 with that estimated with the method of K93. Fortunately, the variation of the scaled energy $\tilde{e}$ for different types of earthqukes is much larger than the uncertainty in the energy estimate so that our estimates with uncertainties of a factor 2 to 3 are still useful.

We applied this method to earthquakes which occurred in southern California for the period of 1995 to the present. Also, we updated the results for the larger earthquakes (1991 Sierra Madre earthquakes (M=5.8), 1992 Joshua Tree earthquake (M=6.4), 1992 Landers earthquake (M=7.3), 1992 Big Bear earthquake (M=6.4), and the 1994 Northridge earthquake (M=6.7)) applying the newly determined station corrections and directivity effects. Most events are larger than M_L=3.7.

The data for smaller earthquakes have been obtained by Abercrombie [1995] using the down-hole (2.5 km deep) seismic data recorded in the Cajon drilling site in southern California [Zoback and Lachenbruch, 1992]. A distinct advantage of using down-hole data is that they are free from the complex free-surface effects and the large attenuation near the recording site. These are the main factors that cause the large uncertainties in the results obtained with surface instruments, especially for small earthquakes. Although only one station was used, the data set covers a fairly large azimuthal range (approximately 150°) so that the effects of radiation pattern and directivity were averaged out. Most events are within relatively short distances, 25 km, and the wave forms exhibit clean impulsive characters. Thus, these observations are considered among the most reliable for small earthquakes.

The scaled energy $\tilde{e} = E_R / M_0$ for small earthquakes is about 10 to 100 times smaller than those for large earthquakes.

Large ($M_w \geq 4.5$) *Earthquakes*

The values of $\tilde{e}$ is about 1 to 2×10^{-4} for large earthquakes. If the static stress drop $\Delta\sigma_s$ is 10 to 100 bars, this result indicates that the dynamic stress drop, $\Delta\sigma_d$, is 65 to 110 bars for large earthquakes (equation 4). Thus, the dynamic stress drop $\Delta\sigma_d$ for large earthquakes is comparable to, or slightly larger than, the static stress drop $\Delta\sigma_s$.

Small (M_w<2) *Earthquakes*

Small earthquakes appear to be less efficient in wave radiation than large earthquakes. Even if we allow for the potentially large uncertainties in energy estimation, this difference appears to be too large to be attributed to experimental errors, and probably reflects the real difference in the rupture dynamics between small and large earthquakes. The transition occurs between M_w = 2.5 and 5. The values of $\tilde{e}$ is 2×10^{-6} for small earthquakes. If the static stress drop $\Delta\sigma_s$ is 10 to 100 bars, this result indicates that the dynamic stress drop, $\Delta\sigma_d$, is 5 to 50 bars for small earthquakes (equation 4).

Thus, despite the large uncertainties in energy estimation, we believe that the ratio, $\tilde{e} = E_R/M_0$, provides an important information on the physical process occuring on the fault plane during seismic rupture.

Interpretation

The wave forms radiated from earthquakes are complex at high frequency, suggesting that microscopic processes on a fault plane are important in controlling the rupture dynamics. Such microscopic processes include frictional melting [Jeffreys, 1942; McKenzie and Brune, 1972; Richards, 1977; Cardwell et al., 1978], fluid pressurization [Sibson, 1973; Lachenbruch, 1980; Mase and Smith, 1985, 1987], acoustic fluidization [Melosh, 1979, 1996], dynamic unloading effects [Schallamach, 1971; Brune et al., 1993; Weertman, 1980; Mora and Place, 1998; Ben-Zion and Andrews, 1998] and geometrical effects [Scott, 1996].

The importance of thermal processes in earthquake mechanics has long been recognized, and a recent study of the deep Bolivian earthquake (M=8.3, depth=637 km) [Kanamori et al., 1998] presented an interesting observational case which suggests a dominant role of thermal processes during faulting. For this earthquake, the released potential energy, 1.4×10^{18} J, is almost 30 times larger than the radiated energy, with a very large amount of non-radiated energy (comparable to the total thermal energy released during the 1980 Mount St. Helens eruption) deposited in a relatively small fault zone over a time scale of less than a minute.

The thermal process during faulting would cause a complex sequence of events including local melting, freezing, fluid pressurization, micro-fracturing and injection of fluids. Although these microscopic processes are important for understanding rupture dynamics, it is difficult to determine how these processes work in detail during faulting, because of the limited resolution of seismic methods. In our approach, we use integrated macroscopic parameters such as M_0 and E_R to investigate this problem.

If we consider a gross thermal budget during faulting under a frictional stress σ_f, then the total heat generated during faulting is $Q=\sigma_f DS$. If we assume that the heat is distributed during seismic faulting within a layer of thickness w around the rupture plane, the average temperature rise ΔT is given by

$$\Delta T=Q/C\rho Sw=\sigma_f D/C\rho w \qquad (7)$$

where C is the specific heat, and ρ is the density. In general D increases with the earthquake magnitude, M_w, or M_0, and we obtain

$$\Delta T = (16/7)^{2/3}(1/\pi)\sigma_f \Delta\sigma_s^{2/3} M_0^{1/3} / \mu C\rho w \qquad (8)$$

We computed ΔT from (8) as a function of magnitude M_w for w=1 mm. We assumed $\Delta\sigma_s$=100 bars, C=1 J/g°C, and ρ=2.6 g/cm^3. If w=1 mm, ΔT exceeds 1000 °C at M_w=5 even for a modest value of friction, σ_f=100 bars. Even for w=1 cm, ΔT exceeds 1000 °C at M_w = 7 for the same value of friction. If σ_f> 100 bars, ΔT exceeds 1000 °C at a lower M_w. Thus, thermal process becomes important for large earthquakes.

Depending on whether fluid exists or not in a fault zone, two distinct thermal processes can happen. If there is no fluid in a fault zone, the temperature can rise to cause frictional melting. If no fluid exists, frictional melting is likely to occur for earthquakes with M_w=5 to 7. This general conclusion appears unavoidable even if the values of $\Delta\sigma_s$, σ_f, and w used in (8) are varied over fairly large, but plausible, ranges.

If fluid exists in a fault zone, fluid pressurization could occur. This concept was introduced to seismology by Sibson [1973], and analyzed in great detail by Lachenbruch

[1980], and Mase and Smith [1985, 1987]. If fluid does not escape (small permeability) and the surrounding rock is not compressive, the pressure increase would be of the order of 10 bars/deg [Lachenbruch, 1980]. In actual fault zones, permeability and compressibility vary and the pressure increase may be less. Although the distribution of permeability can be very complex, pressure fluidization can play an important role, at least locally, in reducing friction. A modest ΔT of 100 to 200° would likely increase the pore pressure, thereby significantly reducing friction. This can occur for earthquakes with M_w =3 to 5. According to Chester and Chester [1998], the internal structure of the Punchbowl fault, California, implies that earthquake ruptures were not only confined to the ultracataclasite layer, but also largely localized to a thin prominent fracture surface. They suggest that mechanisms that are consistent with extreme localization of slip, such as thermal pressurization of pore fluids, are most compatible with their observations.

Since a fault zone is probably very complex and heterogeneous in stress, fluid content, permeability, porosity, and compressibility, no single process is likely to dominate. In other words, we do not necessarily expect a single continuous layer of melting and pressurization; we envision, instead, a fault zone that consists of many micro-faults (subfaults) where different mechanisms are responsible for slip at different stress levels, producing complex rupture patterns as observed.

Our interpretation is that, for large earthquakes, melting and fluid pressurization reduce dynamic friction rapidly thereby causing rapid brittle failure resulting in a relatively large $\tilde{e}$. Since both $\Delta\sigma_s$ and $\Delta\sigma_d$ are of the order of 100 bars, and the friction is low, the entire process must be occurring at a stress level comparable to the static and dynamic stress drops, about 100 bars. This is consistent with the result of Beroza and Zoback [1993] and Zoback and Beroza [1993] who found from the diversity of aftershock mechanisms that the friction during the 1969 Loma Prieta, California, earthquake was very low. Also Spudich [1992] and Spudich et al. [1998] inferred from the rotation of slip vectors that the absolute stress during faulting of several earthquakes is comparable to stress drops.

Qualitatively, if the friction drops rapidly, fault motion would take place rapidly, and more energy will be radiated for a given M_w, and results in large $\tilde{e}$. In contrast, if the friction drops gradually, the fault motion is accelerated slowly thereby radiating less energy than the case for sudden drop in friction.

State of Stress

The result presented above suggests that the stress level along mature faults where large earthquakes occur must be low because of the dominant thermal effects such as frictional melting and fluid pressurization. Because of melting or pressurization, a fault zone is self-organized into a low stress state. That is, even if the stress was high in the early stage of fault evolution, it would eventually settle in a low stress fault, after the occurrence of many large earthquakes This state of stress is consistent with the generally held view that the absence of heat flow anomaly along the San Andreas fault suggests a shear strength of about 200 bars or less [Brune et al., 1969; Lachenbruch and Sass, 1980]. The stress in the crust away from active mature faults can be high as has been shown by many in-situ measurements of stress [McGarr, 1980; Brudy et al., 1997]. The stress difference is large, and a kbar type stress may be involved in small earthquakes, but the events are in general so small that it is hard to determine the stress parameters accurately.

The important thing, though, is that as long as the length of the fault is small, the state of stress in the fault zone would not affect the regional stress drastically. However, as the fault developed to some length, some sort of self-organization occurs and the fault settles at a stress level somewhat higher than that on more active plate boundaries.

Magnitude-Frequency Relationship for Mature Faults

One probable consequence of sudden reduction in friction when slip exceeds a threshold value would be runaway rupture. In this context, an interesting observation is the magnitude-frequency relationship for some mature plate boundaries such as the San Andreas fault. For example, the absence of events with magnitude between 6.5 and 7.5 on the San Andreas fault in southern California, despite the occurrence of magnitude 8 earthquake in 1957 (Fort Tejon earthquake) and the average repeat time of about a few hundred years [Sieh, 1984], has been thought somewhat odd. A magnitude-frequency relation for the San Andreas fault has been reported by Wesnousky [1994]. Earthquakes with M from 6 to 7 appear to be fewer than expected for the conventional magnitude-frequency relationship. This observation can be interpreted in terms of the runaway process discussed above. As the magnitude exceeds a threshold value, about 6.5 for the San Andreas, the friction drops drastically so that fault slip cannot stop until it reaches some limit imposed by the regional seismogenic structure or loading geometry. This is a runaway situation caused by dynamic effects of faulting.

If the specific fracture energy, G^*, is constant, the Griffith type cracks are unstable, *i.e.* if the crack length exceeds a threshold, the crack will runaway. So, in this sense all earthquakes, small and large, can get into runaway rupture. In the actual fault zone, G^* is not constant, and the place where G^* is large acts as a barrier to stop rupture propagation [Aki, 1977]. Then the question is what is the probability of some barries stopping the rupture. The easiest way to look at this problem is to use the stress intensity factor K which is given by $(\sigma_0-\sigma_f)(\pi l)^{1/2}$ for a Mode III crack where l is the crack length. As a fault grows, slip increases and the friction, σ_f, drops and l increases. The combined effect of the decreasing σ_f and increasing l increases K. Since the crack extension force is proportional to K^2, the fault rupture becomes harder to stop and runaway rupture is more likely to occur.

The magnitude-frequency relationship is usually understood as a manifestation of heterogeneity of fault structure [Scholz and Aviles, 1986; Okubo and Aki, 1987; Aviles et al., 1987]. In addition to this static feature, slip-controlled dynamic runaway process could be an important element that determines the earthquake statistics for mature faults.

Seismic Breakaway Phase

In a series of papers, Ellsworth and Beroza [1995, 1998] and Beroza and Ellsworth [1996] showed that the moment rate of many earthquakes is initially low but after some time it grows rapidly. They called this sudden increase in the moment rate a breakaway phase. The breakaway phase could be a manifestation of the slip-controlled runaway rupture. However, our model has highly heterogeneous distribution of strength and would not explain the scaling relation proposed by Ellsworth and Beroza [1995, 1998] and Beroza and Ellsworth [1996]. Similar observations, on various time scales, have been made by Umeda [1990, 1992], Kikuchi [1997] and Ruff [1999].

Ground Motion from Large Earthquakes

The effect of a pulse-like near-field ground motion on large structures is becoming an important engineering problem [Heaton, 1990; Heaton et al., 1995; Hall et al., 1995]. However, very few recordings of near-field ground motion from large earthquakes exist. In modeling studies, the records from small earthquakes are used to estimate ground motions from hypothetical large earthquakes. This is a reasonable approach but the possibility exists that the slip velocity during very large earthquakes could be significantly larger than that for small earthquakes because of the thermally controlled reduction in friction.

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